MAPS OF GROUND MOTION AMPLIFICATION REPRESENTING THE 3D EFFECTS OF THE MISSISSIPPI EMBAYMENT ON WAVE PROPAGATION

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Introduction

In this investigation we proposed to constrain ground motion characteristics within and around the Mississippi embayment and to evaluate the effects of the embayment structure on wave propagation by analyzing the response of irregular basement structure using 3D finite-difference calculations. Our project started in March of 2001 and during this twelve-month period we have collected earthquake data from the waveform data recorded by various networks operating in the Mississippi Embayment to identify the embayment structure effects. There are three primary strong-motion networks which are operated by LDEO (Lamont Doherty Earth Observatory), USGS and University of Kentucky (Figure 1) in addition to other broadband stations operated by CERI (Center of Earthquake Research Institute, Memphis). In the follow-up investigation, we are examining these waveforms from earthquakes occurring in the Mississippi Eambayment to determine how well they can be understood using only 1D models or if the waveforms require station specific crustal models. For earthquakes with known focal mechanisms and depths, we applied the technique of Xu and Wein (1997) to refine the initial 1D model by minimizing the misfit in waveforms of individual stations to delineate variations in the respective crustal models. We are also examining the effect of irregular crustal model by simulating finite-difference seismograms using 2D profiles which extends from Memphis to Saint Louis (Catchings, 1999). We embedded seismic sources at two separate locations and simulated waveforms to examine the features that distinguish the effect when seismic waves propagate from the Mississippi Embayment towards the Illinois basin and vice versa. Meanwhile, we are also collecting the geologic information of the Mississippi Embayment region so that they material properties of shallow structures can be incorporated into our final model. In the following sections, we discuss overall characteristics of strong-motion data, modeling of 2001 Enola

Arkansas, 1991 Risco and 1990 Cape Girardeau earthquakes to probe suitability of 1D models of the Mississippi Embayment and examine the basin effects on the amplitude and duration of the waveforms recorded at wide range of distances.

Strong-Motion and Long Period Broadband Data as Recorded in the Mississippi Embayment

Figure 1 displays the distribution of strong-motions that operated in the Mississippi Embayment. These stations have recorded earthquakes (green circles, right panel Figure 1) which occurred within and around the embayment structure. Figure 2 shows strong-motion waveforms recorded by some stations from an earthquake (09/26/90, mb=4.7, 37.165°N and 89.577°W, h-=12km). Figure 3 shows map of SLU (Saint Louis University), USNSN and IU broadband stations in New Madrid that recorded the 1990 Cape Giradeau, Missouri, 1991 Risco Missouri and 2001 Enola, Arkansas earthquakes. Figure 4 shows a preliminary agreement of the observed data with the synthetics obtained for the Enola earthquake in the frequency band between 0.02 and 0.2 Hz. We inverted long-period data to determine moment-tensor solution and depths, and found the fault plane to have a steep dip angle (about 85°). Clearly many of the stations are nodal as indicated by the strong transverse ground motions. Additional study is needed to extend to high frequency. For paths crossing the embayment we required a low velocity layer at the top of the model and paths to CCM and SLM crossing Ozark did not need low velocity zone. Figure 5 and 6 shows the agreement between data and synthetic seismograms for the Cape Girardeau and Risco events. For these events, our focus was to investigate the crustal models using their known focal parameters. In the following section we discuss how we have carried out investigations to accomplish the goal.

Estimation of Station-Specific 1D Crustal Structure - Method

Following Xu and Wein (1997), we have used difference seismograms constructed using data and synthetic seismograms to form the data vector (Δd) and expressed it by a linear combination of first-order partial derivatives ($\Delta \mathbf{G}$) times their first-order velocity perturbations (Δm) as follows

$$\Delta d = \Delta \mathbf{G} . \Delta m$$

where $\Delta d = d - s(m_0)$, d and $s(mo)$ being the data and its corresponding synthetic estimates obtained from the known source parameters and starting velocity model. The vector $\Delta \mathbf{G}$ is a matrix consisting of partial derivatives, constructed by subtracting $s_n(mo)$ from $s(m_0)$ where $s_n(mo)$ is computed using the same source parameters and the starting model perturbed to velocity perturbation of individual layers by about 1 to 2% separately. Δm is the solution vector estimated using conjugate gradient method.

The problem becomes nonlinear and its success may depend on the proximity of the initial model to the true model. Here $\Delta \mathbf{G}$, Δm and Δd are $m \times 1$, $n \times 1$ and $m \times 1$ matrices, respectively. We only invert for the S-wave velocity because the S-wave and surface waves, which are most sensitive to this velocity, is dominant in regional records. We then use the conjugate gradient method to solve for Δm [*Claerbout*, 1992]. The model perturbations Δm are added to the initial model $m_0^{(old)}$ for the new model $m_0^{(new)}$

$$m_0^{(new)} = \Delta m + m_0^{(old)}$$

The synthetics are computed using a reflectivity code that quickly computes the complete synthetic seismogram. The partial derivatives are computed using simple finite difference. We do not iterate on the residuals but instead recompute the partial derivatives using the new model.

The method usually takes less than 10 iterations and requires a good initial model by a process of trial and error. We can invert multiple components and stations for a single event or multiple events depending upon the structure of the region, which must be near the 1-D model for the frequency band of interest.

Application to 1990 Cape Girardeau, 1991 Risco and 1991 Enola Earthquakes

We applied the above approach to three regionally recorded earthquakes within the Mississippi embayment region and applied the method to regional waveforms recorded at station CCM from the 1990 Cape Girardeau, Missouri, 1991 Risco, Missouri earthquakes. The source parameters for the 1990 and 1991 events were determined by *Langston* [1994] using a grid search scheme and velocity model calibrated from refraction and teleseismic receiver-function inversion. Although the receiver-function approach can resolve deeper structure, it has poor resolution of the regional crustal structure. The source mechanism for the 2001 Enola event was determined using moment tensor inversion of long period waves recorded at 10 stations and the depth using a grid search. With this many stations the long period moment tensor is well constrained but mismatches in phase and amplitude between the data and synthetics at shorter periods indicate that improvements can be made to the model. We inverted single station 3-component data and compared the results obtained for different paths to examine any azimuthal variations in earth structure.

The raw data were processed to displacement and the entire record of all three components were used. The long period p-waves are usually lower in amplitude and in the noise in regional records so we only invert for S-wave velocities. We computed synthetics by perturbing the S-wave velocity for only one layer at a time by about 0.1 km/s. The earth is parameterized as multiple layered medium with depth variations in S-wave velocities. We used a Poisson ratio of 0.25 to relate the P-wave velocities (α) to the S-velocities and assumed a relationship of $\rho = 0.32\ \alpha + 0.77$ to constrain and densities (ρ) in order to limit the number of unknowns. This relation may not be appropriate for the shallow sediment structure. For the Cape Girardeau event we found using synthetic tests that the lower crust was poorly constrained and insensitive to the S-velocity perturbations. We therefore represented the entire crust between 25 and 40 km depths by a single layer to prevent trade-off between the surface layer and lower crustal layer velocities. The layer thicknesses were fixed but the thicknesses could vary with depth depending on the resolution for that layer depth. Low velocity surface layers can also be added to better fit the surface wave train. The data and synthetic waveforms were bandpassed using a 3-pole Butterworth acausal filter between 100 or 50 and 10 or 5 seconds period. We normalized the data and synthetics before the inversion for the vertical and horizontal components to have even weights. The tangential component was weighted strongly when the station receiver azimuth was near nodal.

Table of Events.

Origin Time (UT)	Lat (°N)	Lon (°W)	h (km)	M_w (mag)	ϕ (strike)	δ (dip)	λ (rake)	Location
90/09/02 13:18:51.33	37.17	89.58	15	4.3	59	60	155	Cape Girardeau, Missouri
91/05/0 01:18:54.91	36.56	89.82	8	4.2	88	64	-11	Risco, Missouri
01/05/04 06:42:12.68	3.21	92.19	5	4.5	117	86	-32	Enola, Arkansas

General Inferences on the 1D Crustal Models

The above inversion resulted in an improvement of the arrival time and amplitude of the major phases Sn, Sg, SmS, and Rayleigh waves which were not produced by the initial model. The propagation paths from event 1 and 2 to CCM indicate that the average crustal velocity is between 6.2 and 6.5 km/s with a high velocity layer at 10 km depth and a low velocity zone (LVZ) at 25 km depth. The CCM record for event 2 also resolves a gradual gradient from 5.5 to 7 km/sec from 25 to 40 km with upper mantle velocity of 7.8 km/sec. This is poorly constrained for event 1 although the average velocity for that layer is the same and placing the gradient in the event 1 path neither worsens or improves the fit to the data. These models are in good comparison to the CCM receiver function inversion by *Langston* [1994] although the LVZ and high velocity lid in the upper crust is not present in the receiver function result. Event 2 also traveled more through the embayment relative to event 1 so we add a low velocity surface layer to better fit the synthetics which is not needed in event 1. A lot more is desired to be achieved for fitting the shorter-period waveforms which we expect to accomplish during our on-going studies.

Lateral Crustal Structure

The lateral heterogeneity of the Mississippi embayment structure has recently been delineated (Dart and Wolfs, 1998 and Mooney and Andrews, 1984), based on much drill-hole data from the upper Mississippi Embayment (Dart, 1992) and COCORP deep reflection profiles across the buried Reelfoot rift (Nelson and Zhang, 1991). In the neighboring Wabash Valley region, new evidence on paleoearthquakes has emerged from liquefaction data (Obermier et al., 1993). A special issue of SRL (July/August, 1997) on the Illinois basin contains several articles providing useful data on the basement structure of the WVSZ. We have been compiling the seismic, geologic and geophysical data available from such sources for integrating into a first-order 3D velocity model of the embayment region in the New Madrid extending into the Wabash Valley source region. Our plan here is to constrain ground motion characteristics within and around the Mississippi embayment and evaluate the effects of the embayment structure on wave propagation by analyzing the response of irregular basement structure using 3D finite-difference calculations. The initial structure model is prepared based on the contour maps of the relic structure in the Precambrian basement of the Reelfoot rift (Dart and Wolfs, 1998) which is based on a combination of new detailed depth-to-basement data in the Reelfoot rift (the lower left rectangle in Figure 1) and less detailed basement and outcrop data in the surrounding region. As new useful and critical data become available, we plan to incorporate them and modify the initial model. We also have a preliminary 3D basement map for a portion of the Wabash Valley (the upper right rectangle in Figure 1, Bear *et al.*, 1997) which is also a site of large paleoearthquakes (Obermier et al., 1993).

The principal feature of the Mississippi Embayment crust is the uplift of the deep crustal layers and the thinning of the upper crust. Earlier seismic refraction studies conducted by the U.S. Geological Survey in 1980 in the northern embayment (Mooney *et al*., 1983; Ginzsburg *et al.*, 1983) show the variation of deep crustal structure of the northern Mississippi Embayment along and across the Reelfoot rift. Later discovered features of shallow structure such as the sub basins (SB), and major and minor structural highs that lie within upper and lower intra-rift basins are also crucial for understanding waveguide effects. Our current effort is primarily to assemble all these information together to develop the desired 3D model of the Embayment structure.

<u>Finite-Difference Simulation of Mississippi Embayment Structure</u>

To investigate the variation of seismic waveforms as caused by the variation in the crustal structure, we have simulated finite-difference seismograms using a crustal structure across the New Madrid rift complex (NMRC, Braile et al., 1997), extending from Memphis, Tennessee to the east of Saint Louis, Missouri. The USGS acquired refraction data for this approximately 400 km long profile which was investigated by Catchings (1999) to investigate the regional Vp, Vs, Vp/Vs and Poisson's ratios along the path. This later study yielded an extensive 2nd velocity profile from Memphis to Saint Louis, which is to our delight inclusive of mini basins along the surface for the entire path. The path incorporates the Illinois basin too (Figure 7).

The model discretized in this study extends 400 km along from Memphis to Saint Louis and extends to a depth of about 55 km with a grid spacing of 200 meters. Next, we have used an event with a dip and rake of 70^o and 30^o respectively, a rise time of 2s and a f_{max} of 1.4 Hz to compute finite-difference seismograms with a source buried within the upper crust at a depth of 6 km -- 20 km off of the both ends of the model. For each source, the first receiver is located 30 km from the end and interspersed equally to a distance of 380 km. Anelastic attenuation was represented by Q_β of 400, 500 and 800 for the upper, middle and lower crust, respectively. No distinction was made in Q properties for the uppermost and upper crust. Two possible cases were analyzed. First, we ran using model which includes the middle crust and part of the lower crust, thus ensuring that no seismic energy is reflected back from the moho discontinuity. The second model incorporated some structure below the moho discontinuity.

We used the recent computation finite-difference algorithm and executed on a SUN work station (Graves, 1998; Pitarka, 1999). Figure 8 shows the vertical and radial strong-motion accelerograms for the two cases discussed above for the source located near Memphis. Note the strong phases that are present at receivers from 1 (r=10 km) to 8 (r=80km) which reduce in strength till the receiver 30 (r=320 km). While the accelerograms in Figure 8a is for the case that does not include the moho discontinuity. By comparing these accelerograms with those shown in Figure 8b with the moho discontinuity included, we note that surface-wave duration increases. The amplitude of the short-period body wave motions reflected from the moho also increases as a function of the receiver locations As the model extends from Memphis to Saint Louis, the upper crust starts to thin near from Dexter (New Madrid) toward Saint Louis. Clearly this thinning of the structure seems to extend the duration of the surface waves. However, there is less recognizable differences in accelerograms computed for the source located near the Illinois basin in the middle crust (Figures 9a,b). Accelerograms at receivers 36 (r=10km) to 29 (r=80 km) are not as strong as those in Figure 8 for receiver 1 to 8 where the source was buried within the upper crust and significant waves were reflected from the upper crust discontinuity. We also notice differences in the finite-difference seismograms computed for the two models. The S body waves are stronger over a longer period and total durations of the surface waves are also extended for the receivers within the Memphis region when the moho discontinuity is included. In addition, the attenuation is different for the source located near Illinois basin when remains flat compared to the case when seismic energy propagates from Memphis toward Saint Louis.

Second Year's Work

We have not encountered any unanticipated problems in the course of this project. In pursuance of our second year goals and remaining period of the first year of the contract period, we will continue to investigate the impact of the Mississippi Embayment structure on the waveforms recently recorded by Enola earthquake using 3D model. It will be a point source calculation for a portion of the entire Embayment structure. We will analyze short-period seismograms from the SLU network to further investigate if recorded data show the same attenuation characteristics as shown by the finite-difference calculation above. Following the progress of the first year of this project, we intend to develop source models for large earthquakes in the New Madrid and Wabash Valley seismic zone, simulate 3D long-period ($T \geq 1$s) response and investigate the levels of broadband time histories using methods that we have developed at URS (Saikia, 1997; Somerville et al.,1995).

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